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Citation for published version:

Ó Dochartaigh, BÉ, Archer, NAL, Peskett, L, Macdonald, AM, Black, AR, Auton, CA, Merritt, JE, Gooddy, DC & Bonell, M 2018, 'Geological structure as a control on floodplain groundwater dynamics', *Hydrogeology Journal*. <https://doi.org/10.1007/s10040-018-1885-0>

Digital Object Identifier (DOI):

[10.1007/s10040-018-1885-0](https://doi.org/10.1007/s10040-018-1885-0)

Link:

[Link to publication record in Edinburgh Research Explorer](#)

Document Version:

Publisher's PDF, also known as Version of record

Published In:

Hydrogeology Journal

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Geological structure as a control on floodplain groundwater dynamics

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Received: 29 March 2018 / Accepted: 17 October 2018
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Abstract

Groundwater in upland floodplains has an important function in regulating river flows and controlling the coupling of hillslope runoff with rivers, with complex interaction between surface waters and groundwaters throughout floodplain width and depth. Heterogeneity is a key feature of upland floodplain hydrogeology and influences catchment water flows, but it is difficult to characterise and therefore is often simplified or overlooked. An upland floodplain and adjacent hillslope in the Eddleston catchment, southern Scotland (UK), has been studied through detailed three-dimensional geological characterisation, the monitoring of ten carefully sited piezometers, and analysis of locally collected rainfall and river data. Lateral aquifer heterogeneity produces different patterns of groundwater level fluctuation across the floodplain. Much of the aquifer is strongly hydraulically connected to the river, with rapid groundwater level rise and recession over hours. Near the floodplain edge, however, the aquifer is more strongly coupled with subsurface hillslope inflows, facilitated by highly permeable solifluction deposits in the hillslope–floodplain transition zone. Here, groundwater level rise is slower but high heads can be maintained for weeks, sometimes with artesian conditions, with important implications for drainage and infrastructure development. Vertical heterogeneity in floodplain aquifer properties, to depths of at least 12 m, can create local aquifer compartmentalisation with upward hydraulic gradients, influencing groundwater mixing and hydrogeochemical evolution. Understanding the geological processes controlling aquifer heterogeneity, which are common to formerly glaciated valleys across northern latitudes, provides key insights into the hydrogeology and wider hydrological behaviour of upland floodplains.

Keywords Groundwater/surface-water relations · Floodplain · Hillslope · Hydrochemistry · UK

Electronic supplementary material The online version of this article (<https://doi.org/10.1007/s10040-018-1885-0>) contains supplementary material, which is available to authorized users.

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Introduction

The processes controlling water flow, and in particular subsurface flows, from a hillslope through the floodplain to a river, are still not fully understood, including in meso-scale, upland catchments where most runoff is generated (e.g. Tetzlaff et al. 2014; Blume and Meerveld 2015). Understanding groundwater processes and interaction with surface waters is critical to the ability to predict catchment runoff and water quality responses (e.g. Scheliga et al. 2018), and therefore to predict catchment hydrological dynamics at the resolution needed for environmental, including flood, management. This is of increasing importance in many northern latitudes, including the UK, given anthropogenic catchment modifications and escalating extreme weather events that are changing patterns, amounts and rates of runoff generation (Hannaford and Buys 2012; Pattison and Lane 2012). This paper addresses the influence that geological structure and heterogeneity have on groundwater in a floodplain and adjacent hillslope, floodplain and river. The project

involved collection, interpretation and synthesis of detailed geological, hydrogeochemical, and hydraulic (aquifer properties and piezometry) evidence, to investigate the full lateral extent and depth of a floodplain, including the hillslope–floodplain interface.

Groundwater contributes up to 50% of river flow in UK upland areas (Scheliga et al. 2017), and over recent years there has been increasing interest in its role in floodplain and wider catchment hydrology (e.g. Pitt 2008; Bracken et al. 2013; MacDonald et al. 2014; Tetzlaff et al. 2014), including: floodplain storage (e.g. Zell et al. 2015); floodplain groundwater behaviour during flood events (e.g. Jung et al. 2004); quantifying groundwater discharge to rivers during and between flood events (e.g. Haria and Shand 2006; Marshall et al. 2009); groundwater's role in controlling the timing and duration of runoff and catchment discharge (e.g. Kirchner 2009; Bracken et al. 2013; Tetzlaff et al. 2014); and the strength of coupling between groundwater response and river stage (McDonnell 2003; Seibert et al. 2003; MacDonald et al. 2014). Groundwater in floodplains can also act as a geochemical buffer, and geochemical evolution of groundwaters can influence surface waters, with implications for pollution management (Newman et al. 2006; Pretty et al. 2006; Soulsby et al. 2007).

Floodplain groundwater dynamics, and their role in catchment water movement between hillslopes and rivers, are controlled both by structural conditions (e.g. soil characteristics, floodplain morphology, and the geometry and hydraulic properties of, and interface between, different floodplain and hillslope lithological units) and by driving forces (rainfall, snowmelt and soil moisture), which vary over sub-daily to seasonal scales (e.g. Mouhri et al. 2013; Cloutier et al. 2014; Blume and Meerveld 2015). Head perturbations caused by infiltration of rainfall or rising river stage can cause the propagation of a pressure wave through an aquifer, the speed of which is often referred to as *celerity*; and/or can drive physical groundwater flow that can carry chemical, heat or other tracers, and is measured by *velocity* (e.g. Haria and Shand 2006; McDonnell and Beven 2014).

Upland floodplains in northern latitudes, including much of the UK, have experienced a complex glacial and post-glacial history that has typically resulted in heterogeneous bedded sequences of dominantly coarse-grained sediments with varying proportions of finer-grained sediments, dominantly of glacial, glaciofluvial and alluvial origin, with varying proportions of other sediment types such as peat and lacustrine deposits. Individual sedimentary lithofacies are typically highly variable in thickness and in lateral extent, and this has significant implications for floodplain aquifer permeability (e.g. Ritzi et al. 2000, 2004; MacDonald A et al. 2012).

Many detailed hydrogeological studies of groundwater dynamics in floodplains have been carried out, including on lowland (e.g. Jung et al. 2004; Macdonald D et al. 2012) and upland (e.g. Mattle et al. 2001; Diem et al. 2014) floodplains. Floodplain groundwater is difficult to observe and quantify (e.g. Blume and

Meerveld 2015). It can be costly and logistically difficult, requiring borehole drilling, to collect sufficient direct hydrogeological measurements to confidently characterise the full thickness and extent of floodplain aquifers, their hydraulic properties, groundwater dynamics, and interaction with surface waters. Therefore, studies have tended to concentrate on shallow groundwater to <3 m depth, some focussing on the near-river hyporheic zone (e.g. Boulton et al. 1998; Bencala 2000; Lewandowski et al. 2009; Bradley et al. 2010; Krause et al. 2014; Nützmann et al. 2014; Munz et al. 2017); and some more widely across the floodplain (e.g. Tetzlaff et al. 2014; Scheliga et al. 2018). Many studies usefully apply hydrogeochemical and isotopic techniques to support investigations of floodplain groundwater dynamics, particularly in conjunction with hydraulic and/or numerical modelling approaches, including use of natural tracers (e.g. chloride, dissolved organic carbon and stable isotopes), nutrients (nitrate and phosphate), and residence time tracers such as chloro-fluorocarbons (CFC) and sulphur hexafluoride (SF₆; e.g. Sánchez-Pérez and Trémoières 2003; Fragalà and Parkin 2010; MacDonald et al. 2014; Gooddy et al. 2014).

There is a rich vein of studies using numerical modelling to simulate hydraulic (e.g. McDonnell 2003; Seibert et al. 2003; Ala-aho et al. 2017) and also hydrogeochemical (e.g. Mattle et al. 2001; Schilling et al. 2017) behaviour of groundwater in floodplain systems, and to test different climate or other environmental change scenarios. Such models are based on various levels of observed hydrogeological data, but necessarily involve simplification of the complex reality of three-dimensional (3D) floodplain hydrogeology. This simplification limits their ability to accurately represent local variability in groundwater heads in complex floodplain aquifers (e.g. Mattle et al. 2001; Ala-aho et al. 2017; Schilling et al. 2017).

Effectively observing and understanding subsurface hillslope–river connectivity across floodplains requires a multi-technique approach, including detailed characterisation of the entire 3D and time-variant system (e.g. Blume and Meerveld 2015). This study has systematically collected, interpreted and synthesised geological, geophysical, hydrogeological, hydrological, hydrogeochemical and meteorological data for an upland floodplain aquifer, and developed a high-resolution 3D geological model through which to interpret groundwater dynamics measured in ten carefully sited piezometers. In this way, the influences of geological structure and heterogeneity on groundwater processes and water movement in the hillslope–floodplain–river system are investigated.

Methods

Study site

A combination of hydrometric, hydrogeochemical and geophysical methods was used to investigate geological structure

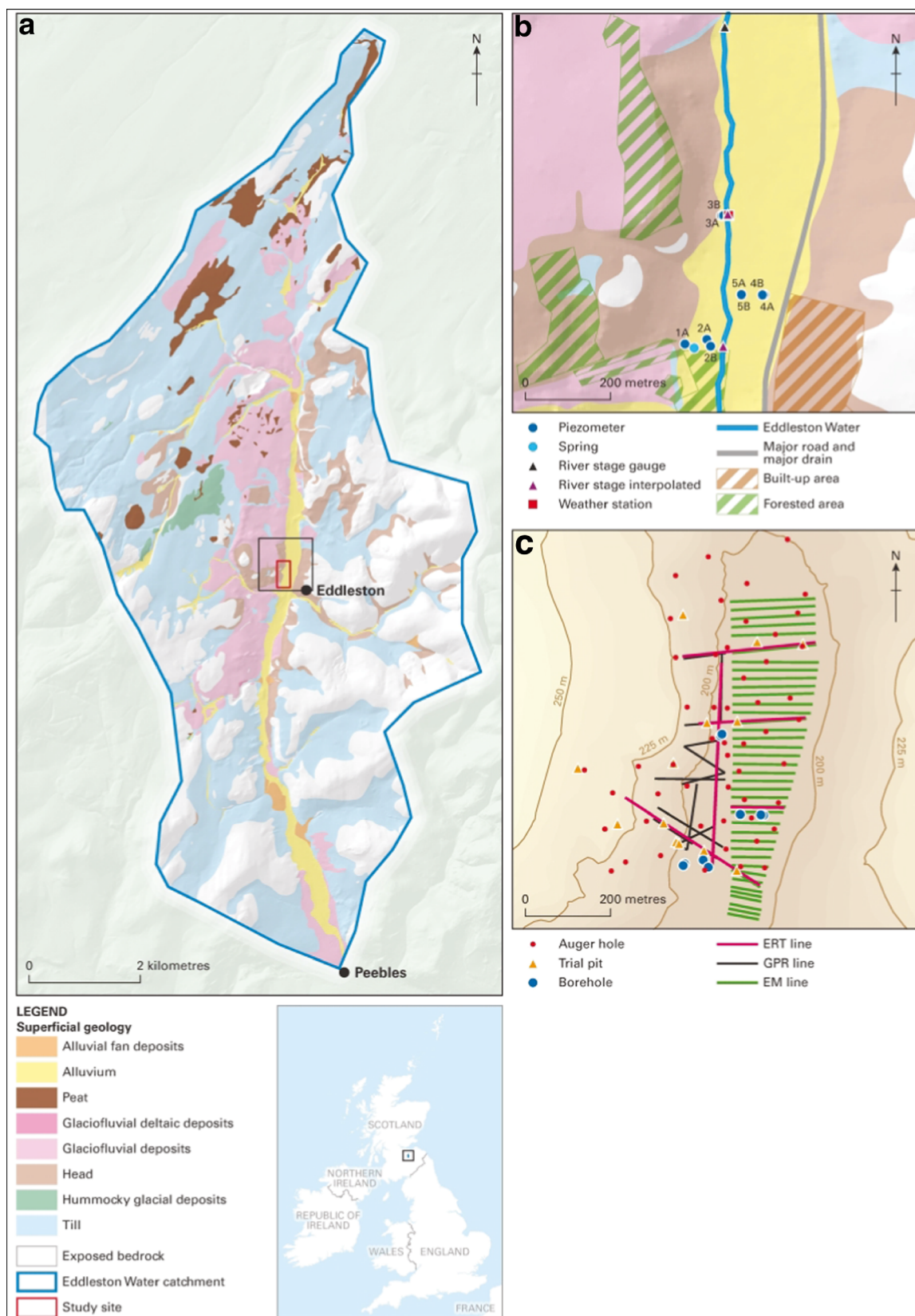


Fig. 1 Location of study site in Eddleston Water catchment in Scotland (UK). **a** Superficial geology of catchment, mapped for this project (© Tweed Forum). **b** Study site showing superficial geology and hydrological monitoring network. **c** Study site showing geological

survey sites, geophysical surface lines and ground surface elevation contours. Groundwater flow directions are shown in the conceptual groundwater model in Fig. 7. Elevation derived from NEXTMap data, supplied under licence from Internap Technologies Inc.

and its influence on groundwater dynamics in a small upland floodplain of the River Eddleston Water (catchment area 69 km²), a tributary of the River Tweed in the Scottish Borders, UK (centre of study site: NGR NT 2425 4755; Fig. 1a). The research is part of a wider project in the Eddleston Water catchment to investigate river restoration options and the effectiveness of natural flood management measures (Werritty et al. 2010; Spray 2016).

The study site area is 0.2 km², stretching ~320 m along the river (Fig. 1b), extending across the floodplain (which is 200–300 m wide in this reach) and partway up the western hillslope with an elevation range of 200–250 m above sea level (asl; Fig. 1c).

Bankfull discharge in the Eddleston Water at the site is estimated at 9.92 m³ s⁻¹ (Werritty et al. 2010) and average flow in 2012 was 0.75 m³ s⁻¹. Estimated average daily rainfall (1990–2009) on the floodplain 1.3 km north of the site at 200 m elevation is 3.78 mm—standard deviation (SD) 5.24 mm—with a maximum daily recorded rainfall of 63.8 mm (Werritty et al. 2010).

The site is typical of many UK upland floodplains. Floodplain land cover is mainly improved grassland, with improved grassland and deciduous and plantation coniferous woodland on the hillslopes. Beyond the study site on the upper hillslopes is extensive heathland. The floodplain is used for spring and autumn grazing, and parts of it for summer silage production. Extensive land use changes have occurred since the eighteenth century, including land drainage, channel straightening, and intensified agriculture (Harrison 2012; Werritty 2006).

Two soil associations dominate: *yarrow* and *alluvium* (Soil Survey of Scotland Staff 1975), also described as *cambisols* and *fluvisols* (IUSS Working Group WRB 2006). Yarrow soils/cambisols occur mainly on the hillslope and are derived from gravels. Floodplain soils are dominated by alluvium/fluvisols and derived from recent silty alluvial sediment with varying amounts of sand and clay (Archer et al. 2013).

Geological, geophysical and hydrogeological surveys

The approach of integrating geological, geophysical and hydrogeological techniques to characterise the 3D floodplain–hillslope environment has been successfully applied in other catchments (e.g. Scheib et al. 2008). A series of geophysical surveys was carried out, comprising 39 electromagnetic induction (EM) lines spaced 20 m apart, 29 ground penetrating radar (GPR) lines, and five two-dimensional (2D) electrical resistivity tomography (ERT; Fig. 1c); whereby EM and GPR penetrate to depths of ~5 m, and were calibrated by trial pit geological logs. ERT penetrates to 20–30 m depth, and was used to target locations for drilling investigation and monitoring boreholes. Geological investigations included field mapping and synthesis of data from previous geological

surveys, digital elevation models (derived from airborne LiDAR), and the geophysical surveys. Shallow intrusive investigations were done by excavating and detailed geological logging of 11 trial pits between 1.1 and 3.85 m deep, and geological logging of a grid of 42 auger holes to approximately 1.2 m depth (Fig. 1c). Deeper geological investigations were undertaken by drilling and geologically logging nine boreholes, which were carefully located to be representative of different parts of the floodplain aquifer system. They comprised four pairs of shallow (<4 m deep) and deep (4.5–8.5 m deep) boreholes, and one single (15 m deep) borehole, along three transects away from the river (Figs. 1b and 2). These were installed with five pairs of shallow (<4 m; suffix B) and deep (4–12 m; suffix A) monitoring piezometers, with two piezometers installed in the deepest borehole (Fig. 2).

Floodplain aquifer properties were determined from constant rate pumping tests on each borehole, varying from 80 to 360 min in duration. Transmissivity was determined using the Jacob approximation for drawdown data and the Theis recovery method (Kruseman and de Ridder 1990).

Hydrological and hydrogeological monitoring

Monitoring was carried out from September 2011 to March 2013. An automatic weather station at the site recorded rainfall at 15-min intervals by a tipping-bucket rain gauge. River stage was monitored at 15-min intervals by gauges at rated sections 400 m upstream and 200 m downstream of the study site (Fig. 1b). River stage at two locations adjacent to the northern and southern piezometer transects was obtained by field surveying of the riverbed datum between the upper and lower gauges (Fig. 1b), and linear interpolation from the gauged values, validated by periodic manual measurement of river stage and corresponding to within ±0.05 m. The five pairs of shallow and deep piezometers (Fig. 1b) were instrumented with pressure transducers to measure floodplain groundwater levels at 15-min intervals. Piezometer pair 1 was sited close to the hillslope/floodplain boundary, pairs 2, 4 and 5 within the main floodplain, and pair 3 in a narrow part of the floodplain close both to the river and hillslope edge.

Three groundwater sampling campaigns were carried out, in October 2011, January 2012 and March 2012, to measure inorganic major, minor and trace ions, dissolved organic carbon, and the groundwater residence time indicator SF₆. All sample analysis was carried out at British Geological Survey laboratories—for details of laboratory analytical methods see Allen et al. (2010) and Goody et al. (2006).

Analytical methods

Geological and geophysical survey data were interpreted and synthesised to develop a high-resolution 3D geological model of the study area, using the GSI3D software package (British

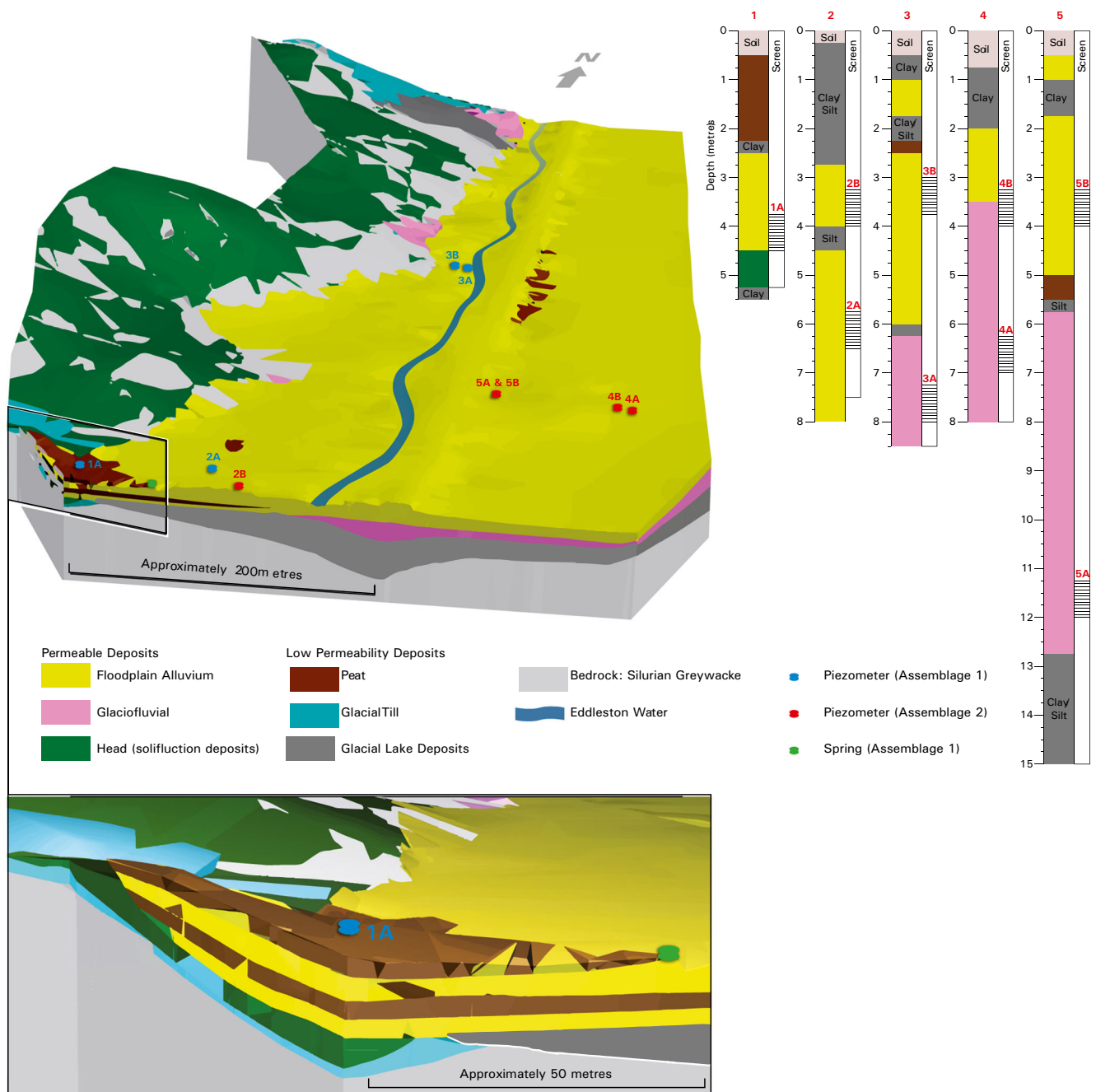


Fig. 2 3D model of superficial geology and location of piezometers and borehole geological logs showing the screened depths of shallow piezometers (2B, 3B, 4B and 5B) and deeper piezometers (1A, 2A, 3A,

4A and 5A). Test pumping was carried out on all piezometers. Clay/silt layers can occur within alluvium, glaciofluvial or head deposits. Insert shows close up of hillslope-floodplain interface. 3D model © Tweed Forum

Geological Survey 2011; Callaghan 2013; INSIGHT Geologische Softwaresysteme GmbH 2018; Ó Dochartaigh et al. 2012). The geophysical (EM, ERT, GPR) and geological data (from trial pit, auger holes and boreholes) were used to construct 67 geological cross sections across the site, which were the basis of the 3D model. The geological model provides a robust foundation from which to interpret the hydrogeological data collected from the ten carefully sited floodplain piezometers.

Hydrogeochemical assemblages for the sampled waters were defined by cluster analysis of selected major ions, using the Ward hierarchical method in the software package R (version 3.0.2) after standardisation of the major ionic values due to the effects of data closure. This was supported by graphical interpretation of base metal ratios and hydrogeochemical parameters.

Relationships between river stage, rainfall and groundwater dynamics in shallow and deep piezometers were

investigated for the period September 2011 to March 2013. The time series data were log-transformed to ensure a normal distribution and cross-correlated using the software package R (version 3.0.2; Chatfield 2004). Mean lag times of peak groundwater level after the onset of rainfall events and the corresponding peak river stage were calculated, to investigate response times of groundwater to rainfall events and river stage changes. Cross-correlation coefficients were plotted with lower and upper 95% confidence intervals ($2/\sqrt{n}$; n = sample size = 9,206 data points) to reveal significant correlations, and response times (lags) were determined as the highest correlated points from each plot.

Results

Floodplain–hillslope geological structure

The floodplain geological structure is highly heterogeneous, comprising a variably thick sequence of unconsolidated superficial deposits of Quaternary age infilling a glacially eroded bedrock valley (Fig. 2). Most of the floodplain is capped by a layer of silt and/or clay, 0.5–2 m thick, interpreted as overbank alluvial deposits. Below this is a layer dominated by alluvial sand and sandy gravel, to 4–8 m depth, containing lenses of silt, clay and peat. In the floodplain centre this overlies a layer of glaciofluvial sand and gravel 4–8 m thick (at a depth of 8–13 m), with discontinuous intervening lenses of clay and peat. The alluvial and glaciofluvial sands and gravels together form a significant aquifer. This is underlain across much of the floodplain by a low resistivity layer of glaciolacustrine silts and clays 10–20 m thick, indicating that a significant glacial lake developed in the valley during its deglaciation. This contradicts previous work by Sissons (1958, 1967), who in the absence of borehole evidence argued against the presence of such a lake.

Rockhead below the floodplain was not penetrated by the boreholes, but from geophysical data is inferred to range from <5 to >25 m depth. Bedrock is expected to be the same as exposed on adjacent hillslopes and across the Eddleston Water catchment (Auton 2011), and to comprise greywacke (well-cemented, poorly sorted sandstone) of Silurian (Palaeozoic) age.

Across much of the adjacent hillslope the uppermost 0.2–0.4 m of the greywacke bedrock is weathered and mostly overlain by thin unconsolidated gravelly head deposits (solifluction deposits derived from bedrock weathering), with minor thin outcrops of glacial till (Archer et al. 2013; Figs. 1 and 2). At the interface of the hillslope and floodplain, head deposits are overlain by, and interlayered with, alluvial sand and gravel deposits, with additional discontinuous but significant interlayering of peat (Fig. 2). The geological

Table 1 Summary of lithology and aquifer properties of floodplain piezometers. Data © Tweed Forum

| Piezometer | Screened | Lithology of | Interpreted geological unit | Rest water level | Artesian | Test yield | Specific capacity | Transmissivity | Hydrogeochemical | |
|------------|----------------|------------------|-----------------------------|---------------------------|---------------------|-------------------------------------|---|-------------------------------------|------------------|---|
| depth (m) | section (mbgl) | screened section | of screened section | before test (mbgl) | conditions observed | (m ³ day ⁻¹) | (m ³ day ⁻¹ m ⁻¹) | (m ² day ⁻¹) | assemblage | |
| 1A | 5.31 | 3.8–4.56 | Sandy gravel | Alluvium or glaciofluvial | –0.03 | Yes | 159 | 873.63 | 1,000 | 1 |
| 2A | 7.61 | 5.83–6.59 | Very sandy gravel | Alluvium | 0.3 | Yes | 13.5 | 135.00 | 220 | 1 |
| 2B | 4.2 | 3.32–3.92 | Very sandy gravel | Alluvium | 0.47 | Yes | 11.8 | 11.46 | 50 | 2 |
| 3A | 8.58 | 7.33–8.09 | Sandy gravel | Glaciofluvial | 0.74 | No | 1478 | 3,016.33 | 415 | 1 |
| 3B | 4.75 | 2.99–3.75 | Sandy gravel | Alluvium | 1.0 | No | 132 | 173.68 | 400 | 1 |
| 4A | 8.02 | 6.25–7.01 | Sandy gravel | Glaciofluvial | 0.55 | Yes | 12.6 | 91.97 | 400 | 2 |
| 4B | 5.04 | 3.28–4.04 | Very sandy gravel | Alluvium | 0.51 | Yes | 12.3 | 82.00 | 200 | 2 |
| 5A | 13.07 | 11.27–12.07 | Gravel | Glaciofluvial | 0.865 | No | 11.5 | 47.92 | 100–200 | 2 |
| 5B | 4.7 | 3.24–4.00 | Sandy gravel | Alluvium | 0.675 | No | 171 | 137.35 | 250 | 2 |

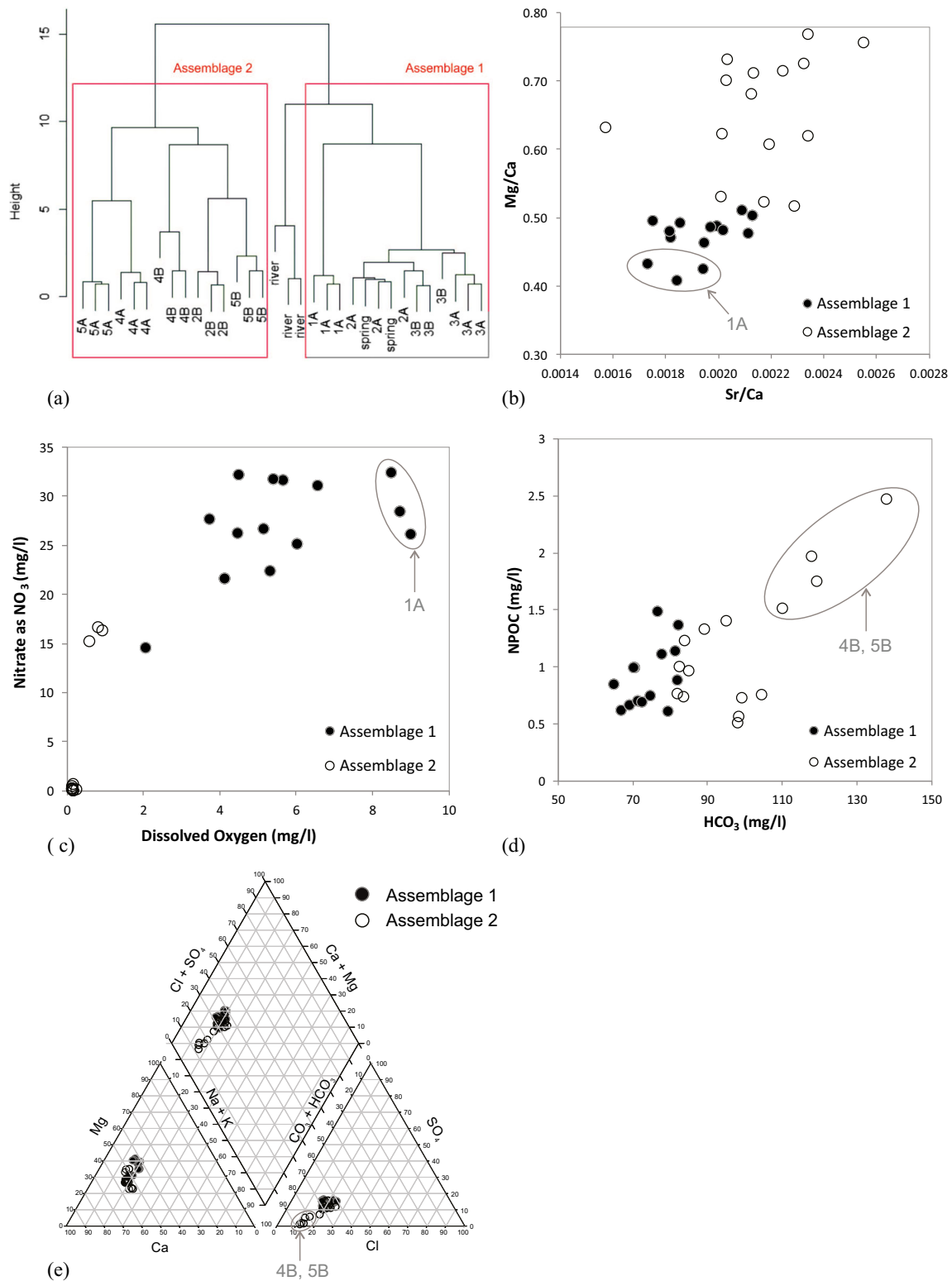


Fig. 3 Selected base metal ratios, and ionic and hydrogeochemical relationships in floodplain groundwaters, and statistical analysis of selected major ion chemistry. **a** Dendrogram obtained by hierarchical cluster analysis using selected standardised major ion chemistry; **b** molar ratios of Sr/Ca and Mg/Ca; **c** dissolved assemblages are distinguished: oxygen and nitrate; **d** bicarbonate and non-purgeable organic carbon (NPOC). Two hydrogeochemical assemblages are

distinguished: assemblages 1 and 2 (see text for details); **e** trilinear (Piper) diagram showing major ion type. Sampled waters labelled by piezometer ID and divided into hydrogeochemical assemblage 1 and assemblage 2 (see text for details). Height on the Y axis of plot (see **a**) is related to the degree of dissimilarity between clusters. All chemistry data were derived from three sampling rounds, in October 2011, January 2012 and March 2012. Data © Tweed Forum

heterogeneity in this interface zone is greater than anywhere else across the study site.

Floodplain and hillslope hydrogeological properties

The alluvial–glaciofluvial floodplain aquifer has a moderate to high transmissivity of generally $200\text{--}400\text{ m}^2\text{ day}^{-1}$ (Table 1). Rarely, transmissivity is as low as $50\text{ m}^2\text{ day}^{-1}$, which is likely to be related to the local presence of low-permeability silt and/or peat lenses in the alluvium. The aquifer in the hillslope–floodplain interface zone shows a high transmissivity of at least $1,000\text{ m}^2\text{ day}^{-1}$, almost certainly related to the interfingering of floodplain alluvium with coarse-grained head deposits. This is equivalent to hydraulic conductivity values of $30\text{--}100\text{ m day}^{-1}$ for the alluvial and glaciofluvial sands and gravels, and up to 500 m day^{-1} for mixed alluvial and head deposits. Hydraulic properties of the basal glaciolacustrine sediments were not directly tested, but glaciolacustrine silts and clays elsewhere in Scotland typically have low permeability (Lewis et al. 2006; MacDonald et al. 2005, 2012): for silts typically $10^{-3}\text{--}10^{-1}\text{ m day}^{-1}$ and for clays typically $5 \times 10^{-7}\text{--}10^{-3}\text{ m day}^{-1}$ (Lewis et al. 2006).

Bedrock transmissivity at the site was not directly measured, but Silurian greywacke aquifers elsewhere in southern Scotland have low productivity (Ó Dochartaigh et al. 2015) with an estimated average transmissivity of $\sim 20\text{ m}^2\text{ day}^{-1}$ (Graham et al. 2009).

A previous study determined field saturated hydraulic conductivity (K_{fs}) of soils across the site at depths of $0.04\text{--}0.15\text{ m}$ and $0.15\text{--}0.25\text{ m}$ (Archer et al. 2013). Floodplain soils under grazed grassland have low K_{fs} (median 1 mm h^{-1}), linked to a combination of soil compaction from grazing animals, a horizon of low-permeability silt/clay in the underlying Quaternary alluvium, and a lack of coarse plant roots that provide preferential flow pathways (Archer et al. 2013). On the western hillslope, the K_{fs} of soils overlying head deposits in areas of grassland and ~ 50 -year-old plantation forest is relatively high ($11\text{--}100\text{ mm h}^{-1}$), and in a mature woodland on the upper hillslope is very high ($> 500\text{ mm h}^{-1}$; Archer et al. 2013).

Floodplain groundwater hydrogeochemistry and age

Two hydrogeochemical assemblages in the floodplain groundwaters (Fig. 2) are distinguished by cluster analysis of major ion chemistry (Fig. 3a), supported by hydrogeochemical parameters and base metal ratios—Fig. 3b–e and Table S1 of the electronic supplementary material (ESM).

Assemblage 1 is characterised by oxygenated groundwater; lower base metal ratios and general lower levels of mineralisation, including lower bicarbonate and typically lower dissolved organic carbon; but notably higher nitrate concentrations than in assemblage 2 (Fig. 3). Assemblage 1 was seen only in the study area in the western side of the floodplain, in

piezometers 1A, 2A, 3A and 3B (Fig. 2). Groundwater closest to the hillslope in piezometer 1A shows the highest dissolved oxygen and lowest base metal ratios (Fig. 3).

Assemblage 2 is characterised by low-oxygen, and usually reducing, groundwater; higher base metal ratios and higher levels of mineralisation overall, including higher bicarbonate; low or negligible nitrate concentrations; and in most cases higher dissolved organic carbon than assemblage 1 (Fig. 3). It was seen mainly on the eastern floodplain in the study area, in piezometers 4A, 4B, 5A and 5B, and in a single shallow piezometer (2B) on the west bank (Fig. 2), which shows very different groundwater chemistry than its deeper paired piezometer 2A. The low concentration or absence of nitrate, combined with reducing conditions—in stark contrast to assemblage 1, and despite this part of the floodplain receiving annual inputs of nitrogen through agricultural slurry spreading—are consistent with nitrate reduction. Higher base metal ratios in assemblage 2 groundwaters suggest they have experienced more water–sediment interaction than assemblage 1 groundwaters, which may indicate long aquifer residence times (Fig. 3).

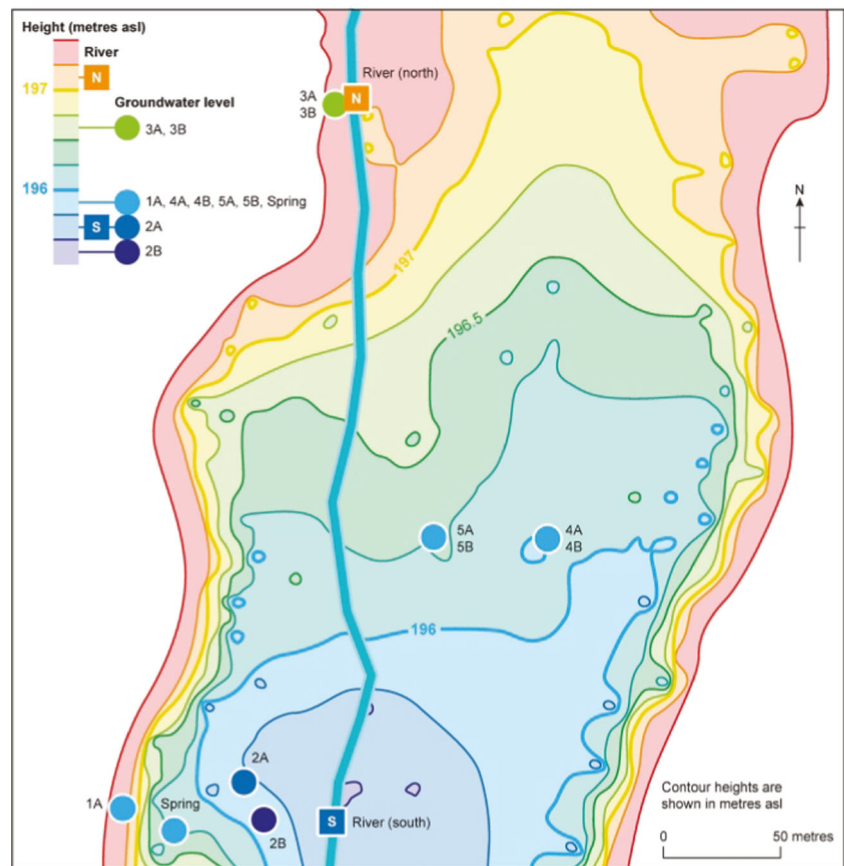
Additional evidence for mean groundwater residence time was obtained from dissolved SF_6 concentrations, which indicate the proportion of modern water in groundwater. Both assemblages contain a proportion of groundwater with mean residence times in the aquifer of at least 20–30 years, but both also contain fluctuating fractions of modern water at different times of year, indicating there are event-scale or seasonal variations in groundwater inflow to the aquifer (Table S1 of the ESM). Overall, groundwaters in both Assemblages contain similar fractions of modern water (assemblage 1: 13–56%; mean 37%; assemblage 2: 15–45%; mean 30%).

Groundwater dynamics

Piezometric groundwater heads are highest upstream and lowest downstream across the site, with a hydraulic gradient of ~ 0.004 (Figs. 4 and 5). Under most conditions, river stage is higher than immediately adjacent groundwater levels (Figs. 4, 5 and 6), causing a hydraulic gradient away from the river that drives water flow from the river into the aquifer. This gradient is reversed following large rainfall events and/or extended wet periods, when groundwater heads recess more slowly than river stage, driving water flow from the floodplain aquifer into the river (e.g. piezometers 3A and 3B in Fig. 6b).

By contrast, at the south of the study site, groundwater head at the hillslope edge of the floodplain in piezometer 1A is consistently higher than heads closer to the river and higher than river stage (Fig. 5b), creating a hydraulic gradient from the base of the hillslope into the floodplain throughout the year. Groundwater level rises correlate with both river stage rises and rainfall events, but across the floodplain for at least 100 m distance from the river there is a significantly stronger correlation between groundwater levels and river stage (mean

Fig. 4 Ground elevation, floodplain groundwater levels and river stage across the study site during a dry period (16/02/2012), illustrating depth to groundwater below ground surface and head gradients. Ground surface elevation derived from NEXTMap data, supplied under licence from Intermap Technologies Inc



cross-correlation coefficient 0.76) than between groundwater levels and rainfall (mean cross-correlation coefficient 0.19; Table 2). The mean cross-correlation coefficient between river stage rise and rainfall is 0.378.

River stage rise is lagged behind rainfall, with a mean lag time of 6 h. Groundwater level rise is lagged behind both rainfall and river stage rise, but with significant lateral variation across the floodplain in response to river stage rise. There are two main patterns:

1. Groundwater-level rises near the centre of the floodplain (2A, 2B, 4B, 5A and 5B) lag closely behind river stage, by 0.93–2.75 h (Table 2; Figs. 5 and 6)
2. Groundwater-level rises near the floodplain edge (piezometers 1A, 3A and 3B) lag significantly longer behind river stage, by 33.75–47.5 h (Table 2; Figs. 5 and 6).

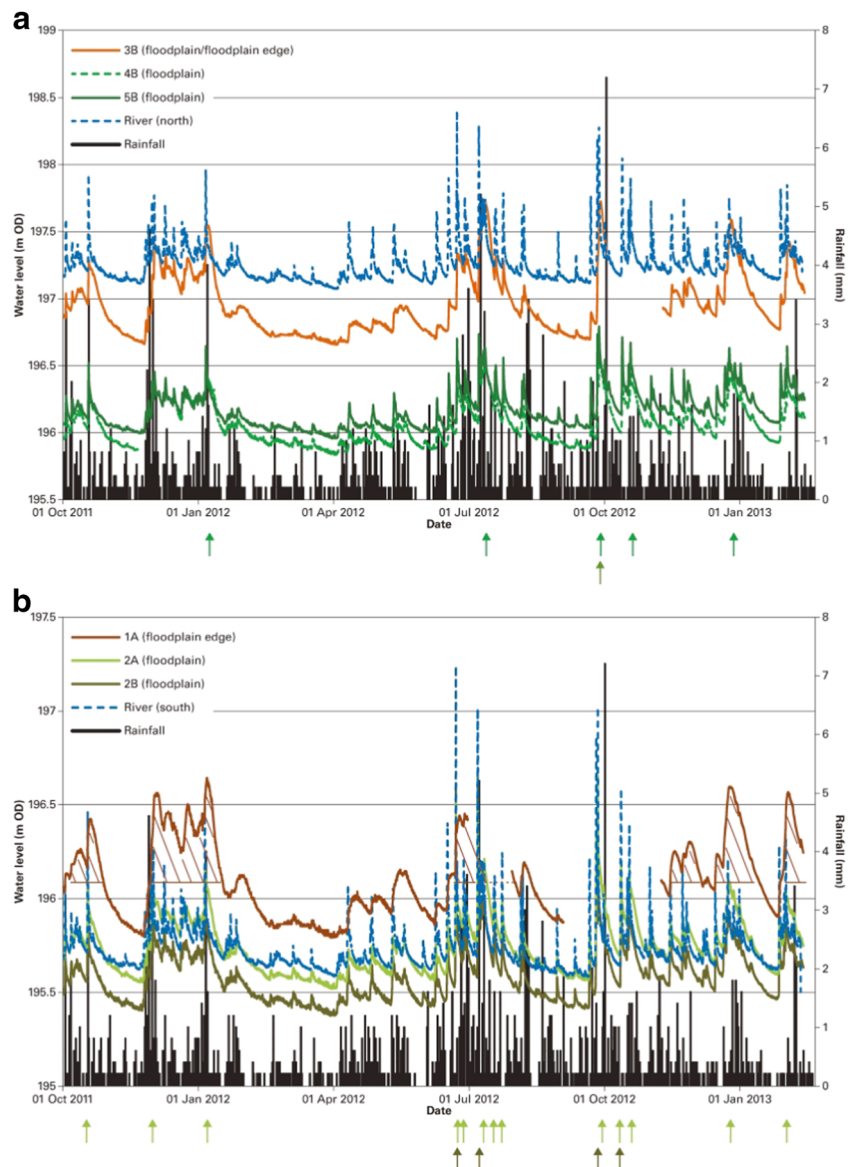
These two patterns are also reflected in a difference in the rate of groundwater level recession. In much of the floodplain, groundwater level recession typically occurs within <1 day, such as in piezometers 2A, 4B and 5B during two selected rainfall events (Fig. 6). Near the floodplain edge, groundwater recession rates are slower, occurring over days to weeks, such as in piezometers 1A, 3A and 3B in the same two rainfall events (Fig. 6).

Groundwater level responses also vary according to depth in the floodplain aquifer. A number of piezometers, at depths of 5–8 m, show evidence of upward groundwater head gradients (e.g. piezometers 2A and 2B: Fig. 5b; piezometers 3A and 3B: Fig. 6b), indicating upward groundwater flow. There is no significant relationship between piezometer depth and groundwater-level/river-stage lag time or correlation function (Table 2). Therefore, although there are local vertical head variations within the floodplain aquifer, there is no evidence of continuous vertical separation of aquifer units.

All the piezometers that show upward head gradients also show temporary artesian groundwater levels, generally preceded by high rainfall and river stage events (Table 1; Figs. 5 and 6). The average duration of artesian conditions is 1–4 days; and rarely up to 8 days (e.g. Figs. 5b and 6b). By contrast, piezometer 1A, at the edge of the floodplain, shows artesian conditions for significantly longer time periods (average duration 20 days; maximum 46 days) (e.g. Figs. 5b and 6a).

Piezometer 2B shows confined conditions. The shallow aquifer here has low transmissivity (Table 1) and is overlain by a particularly thick layer of clay and silt (Fig. 2). Piezometer 2B therefore appears to be in a zone which is relatively isolated from the rest of the aquifer, restricting active inflow from both river and hillslope.

Fig. 5 Groundwater levels, river stage and rainfall for October 2011 to December 2012: **a** in the north and centre of the study site (note that piezometer 4B is 180 m down-valley from 3B and River (north) gauge); and **b** in the south of the study site. Groundwater levels in floodplain piezometers shown in green; groundwater levels in hillslope/floodplain edge piezometers shown in orange or brown. Extended artesian periods in piezometer 1A shown as shaded areas; short artesian periods in 2A, 2B, 4B and 5B shown by arrows below graphs. Data © Tweed Forum

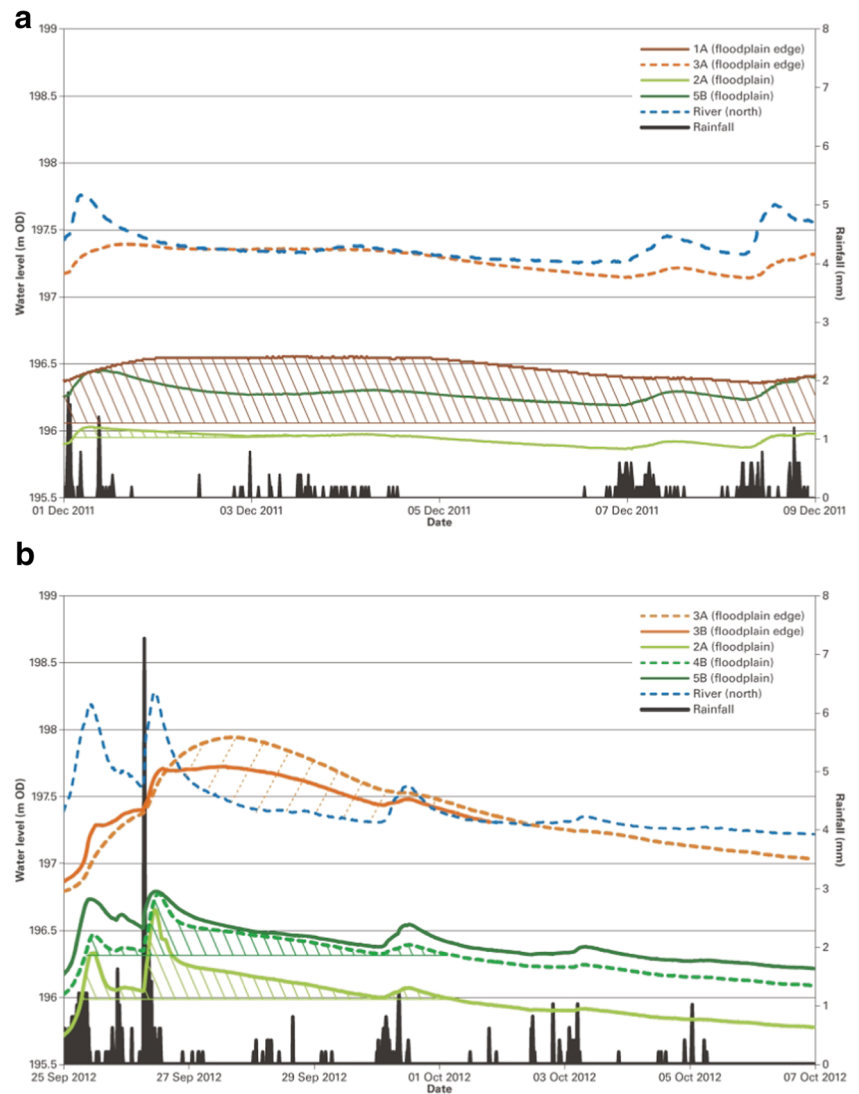


Discussion

At a broad scale, the Eddlestone floodplain aquifer is dominantly permeable and unconfined. Piezometric evidence shows the dominant groundwater flow direction through the floodplain aquifer is down-valley, with local flow directions from the river to the adjacent aquifer, and from the hillslope edge of the floodplain towards the river (Fig. 7). Groundwater can be resident in the aquifer for decades, indicated by SF₆ concentrations and significant hydrogeochemical evolution in some areas. However, there is significant local-scale lateral and vertical heterogeneity in aquifer properties across the floodplain and in the hillslope–floodplain interface, which strongly influences groundwater dynamics, including the timing and duration of artesian conditions, and hydrogeochemical evolution. The aquifer heterogeneity is not random, but is a function of

the detailed glacial and post-glacial Holocene erosion and sedimentation history of the floodplain and hillslope environment—a history that is shared by previously glaciated floodplains in northern UK (e.g. Bell 2005; Scheib et al. 2008) and across northern latitudes (e.g. Bennett and Glasser 2011). The high permeability sands and gravels that dominate the Eddlestone floodplain were deposited initially by glacial meltwater rivers over low permeability glaciolacustrine silts and clays, and later by active Holocene river channels. Lower permeability alluvial silts and clays were deposited in slow flowing abandoned channel meanders or oxbow lakes, or by overbank flooding. Permeable hillslope solifluction deposits and underlying weathered bedrock are also the result of glacial and Holocene erosive and sedimentation processes. Understanding these geological processes is a key step in characterising the detailed geological heterogeneity that has

Fig. 6 Detail of groundwater levels, river stage and rainfall for two selected periods covering rainfall events at the study site: **a** December 2011 and **b** September 2012. Groundwater levels in floodplain piezometers shown in green; groundwater levels in floodplain edge piezometers shown in orange or brown. Shaded areas: solid lines show times when groundwater in that piezometer becomes artesian; dashed lines highlight times when groundwater level in that piezometer rises higher than adjacent river stage. Data © Tweed Forum



such a strong influence on hydrogeological and wider hydrological behaviour in the floodplain.

Lateral geological heterogeneity in the floodplain is evident in distinctly different patterns of groundwater level fluctuation and hydrogeochemistry across the aquifer. The strong correlation between river stage and groundwater level across much of the floodplain, to at least 100 m distance from the river, is driven by changes in river stage, which cause rapid (<3 h) response times in floodplain groundwater levels, corresponding with observed rates of pressure wave propagation through floodplain aquifers reported by authors such as Cloutier et al. (2014), Jung et al. (2004) and Wenninger et al. (2004). By contrast, significantly slower groundwater level rises and recessions in the hillslope–floodplain interface zone are driven by inflow of hillslope water to the floodplain aquifer from the infiltration of local rainfall to permeable hillslope soils. Highly permeable hillslope solifluction deposits facilitate the subsurface transfer of hillslope water into the floodplain aquifer,

which both reduces shallow runoff directly into the river system, and can raise groundwater levels at the edge of the floodplain for several weeks. This response was observed independently of the width of the floodplain (e.g., in piezometers 1A, 3A and 3B), but persisted for longer where the floodplain was wider. There are also lateral variations in hydrogeochemistry, with more geochemically evolved groundwaters seen almost exclusively in the wider eastern side of the floodplain, further from the river and hillslope, where the dominant inflows may be of longer-resident groundwater from up-valley. The only exception to this was in a relatively low permeability, laterally and vertically isolated zone in the western floodplain (at piezometer 2B). In the rest of the western floodplain, the evidence that groundwaters are less geochemically evolved indicates more active recharge and groundwater mixing.

The effect of vertical heterogeneity in the floodplain to a depth of at least 12 m, due to the presence of discontinuous lenses of low permeability clays, silts and peats within the

Table 2 Mean groundwater level lag times in response to rainfall and river stage change at study site, September 2011 to March 2013

| Piezometer | Mean lag of peak groundwater level after onset of rainfall event (hours) | Cross-correlation coefficient | Mean lag of peak groundwater level after peak river stage (hours) | Cross-correlation coefficient |
|------------|--|-------------------------------|---|-------------------------------|
| 1A | 117.25 | 0.125 | 47.5 | 0.700 |
| 2A | 11.5 | 0.194 | 1.5 | 0.808 |
| 2B | 7.25 | 0.237 | 0.93 | 0.878 |
| 3A | 55.75 | 0.167 | 33.75 | 0.691 |
| 3B | 36.75 | 0.148 | 41 | 0.576 |
| 4A | 11.25 | 0.237 | 2.75 | 0.793 |
| 4B | 13.5 | 0.188 | 2.75 | 0.784 |
| 5A | 12.25 | 0.187 | 2.75 | 0.787 |
| 5B | 10.25 | 0.250 | 2 | 0.835 |

A cross correlation coefficient close to 1 denotes high correlation.

dominantly permeable sand and gravel aquifer, is to locally compartmentalise the aquifer. This leads to significant variability in hydrogeochemical evolution, despite similar mean groundwater residence times. Upward hydraulic gradients from the deeper (4.5–12 m) to the shallower (<4 m) aquifer can occur in these zones, in some cases causing artesian conditions, which may lead to groundwater flooding. Restricted groundwater inflows in some zones have also promoted geochemical evolution to less oxygenated, more mineralised groundwaters, in some cases also causing denitrification.

Understanding upland floodplain aquifer heterogeneity and its controls has many benefits—for example, it enables the identification of floodplain zones that are likely to be at greater risk of groundwater flooding, and better estimations of the likely duration of any flooding. Groundwater flooding, driven by artesian conditions, is of growing

concern in floodplains; it can persist for extended periods and have significant impact (e.g. Macdonald D et al. 2012; MacDonald et al. 2014). It also enables the targeting of hydrological monitoring such as observation piezometers, to representative locations, promoting more effective and efficient data collection. Understanding the patterns and scales of heterogeneity in floodplain aquifers and groundwater behaviour may also help the development of more representative numerical groundwater flow models, allowing more realistic characterisation of aquifer structure, properties and boundary conditions.

Conclusions

The Eddleston Water floodplain aquifer, although relatively small, shows significant variability in groundwater flow dynamics and hydrogeochemistry both laterally and with depth across the floodplain and hillslope–floodplain interface. Groundwater levels respond strongly to river stage for at least 100 m distance from the river, rising and falling within hours. By contrast, in the narrow floodplain–hillslope interface, groundwater levels respond more slowly, continuing to rise for days, and can maintain higher water tables for weeks after rainfall events, sustained in part by subsurface inflow from the hillslope.

Geology (lithology and structure) is a key control on this variability, and consequently on the role of groundwater in regulating hillslope–river hydrological coupling. The aquifer comprises permeable sands and gravels locally interbedded with silts and clays, and is strongly linked physically and hydraulically to the hillslope through permeable solifluction deposits. The geological structure of the hillslope–floodplain interface zone is particularly important in controlling water transfer from hillslope to floodplain. The geological heterogeneity is not random, but is a function of the geological processes that have operated throughout the glacial and post-glacial history of this

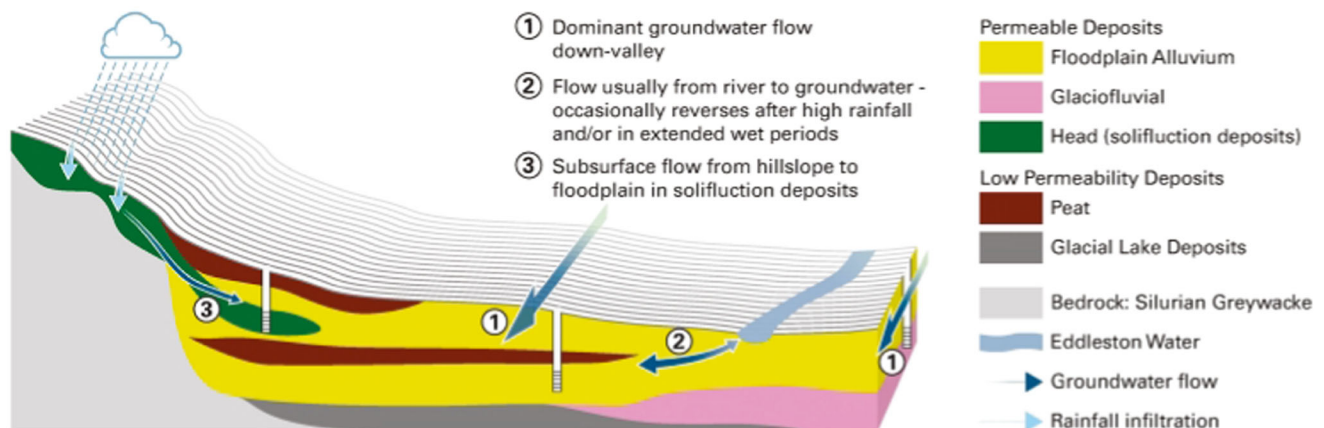


Fig. 7 Conceptual model of groundwater flow in the floodplain aquifer. © Tweed Forum

upland environment. These processes are shared by formerly glaciated catchments across northern latitudes (e.g. Bennett and Glasser 2011). Capturing and accurately representing the detailed structural and lithological heterogeneity of an individual floodplain requires detailed 3D geological and hydrogeological data collection and interpretation. However, understanding the geological processes that created them enables much better initial characterisation of floodplain aquifer structure and properties, and more effective targeting of field investigations to generate necessary new data. An in-depth understanding of geological structure is, therefore, critical to identifying, understanding and predicting groundwater dynamics and hydrogeochemistry, and wider hydrological behaviour, in upland floodplains (for locations of piezometers see Figs. 1 and 2).

Acknowledgements This research is part of the Eddleston Water Project, a partnership initiative led by Tweed Forum with the Scottish Government, Scottish Environment Protection Agency, University of Dundee, British Geological Survey, Scottish Borders Council and others. We thank all partners for their support. We thank Dan Lapworth for support with cluster analysis for hydrogeochemical typology, Craig Woodward for help with figure production, and British Geological Survey laboratory staff for sample analysis. Mike Bonell made a significant contribution to the design and completion of this research, and authored part of an early version of this paper before his passing in July 2014. This paper is published with the permission of the Executive Director of the British Geological Survey (NERC).

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